GRAIL refinements to lunar seismic structure

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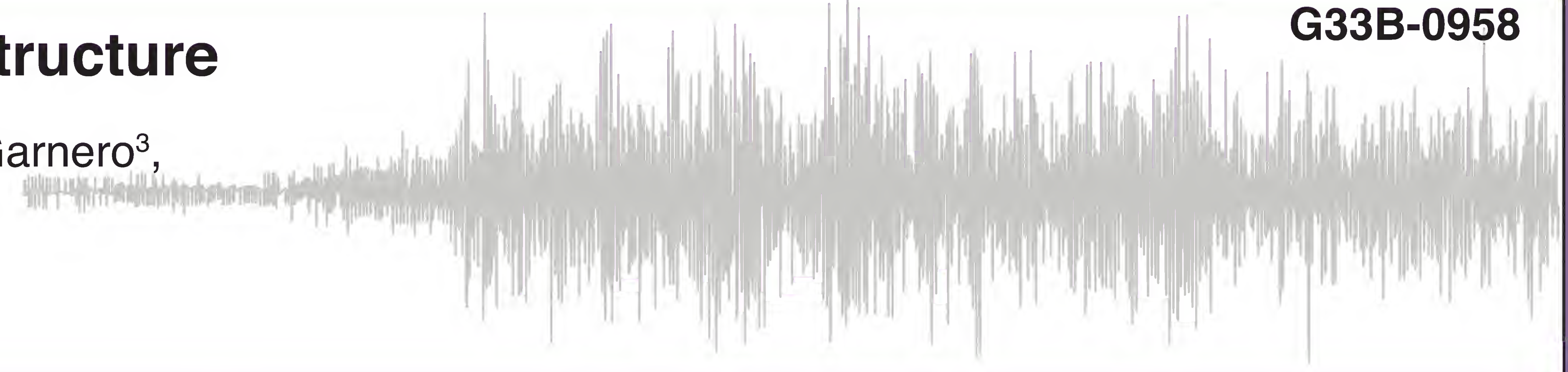

Refinements to array processing of Apollo seismic data

A method to enhance and detect subtle seismic arrivals, typically used in terrestrial seismology, is to stack seismograms that have been time-shifted to the predicted arrival time of a hypothetical phase of interest. We previously applied this array processing approach to the Apollo lunar seismic data [1], providing the first direct constraint on the size and state of the Moon's core (Figure 1). The method used travel time predictions made from pre-existing estimates of crust and mantle velocities and densities and assumed that each of the Moon's layers is a uniform shell, with no lateral variation or heterogeneity. In reality, the structural properties of the Moon are likely inhomogeneous, and vary both laterally and with depth.

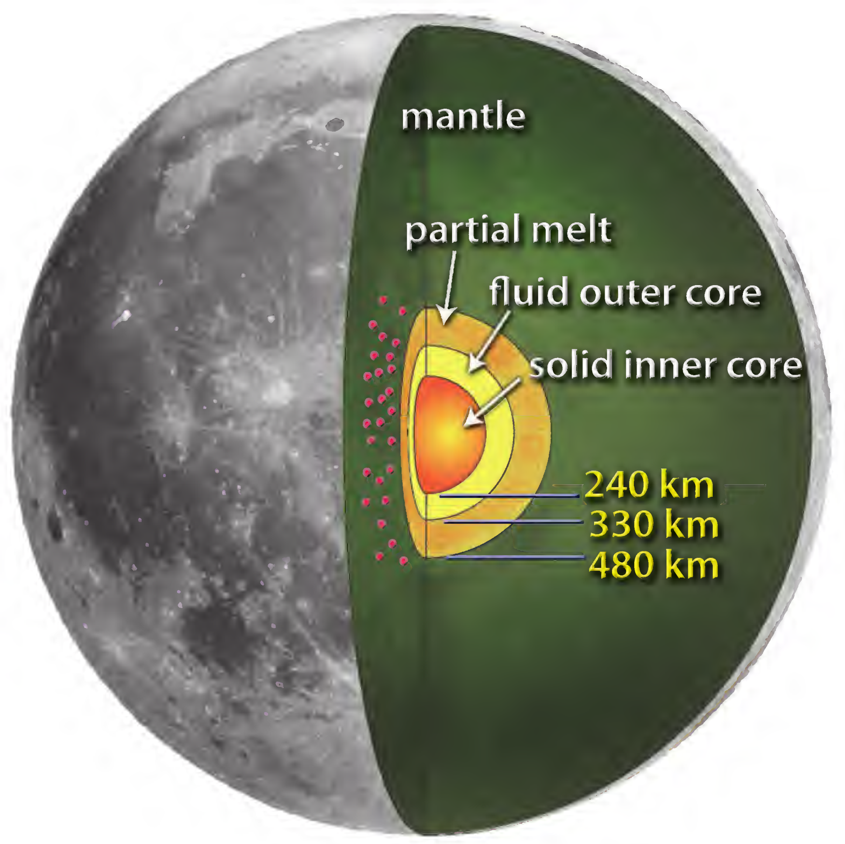

Figure 1: Graphical representation of the seismic velocity model for the lunar core, as constrained in [1]. We searched for core reflections by summing stacks of deep moonquake waveforms along the predicted arrival times of core phases (for example, PcP, a down-going P-wave that reflects off the core-mantle boundary) for a range of hypothetical core radii and layer velocities.

We will adjust the observed times of core-reflected seismic phases from the known distribution of lunar seismic events by including travel-time perturbations based on the following predictions:

1) Refined estimates of crustal thickness derived from GRAIL's gravity model,
2) Variations in mantle velocities based on a suite of velocity models (depth heterogeneity) and seismic tomography (lateral heterogeneity), and
3) GRAIL's constraint on the core radius, layering, and state (solid vs. molten).

For a given ray path generated by a 1D ray-tracer, we will collect the predicted travel time variation from a single model perturbation along that ray path. This process will be repeated iteratively to account for the three perturbations we wish to include. The end result is a total travel time anomaly for the input ray path. For each deep moonquake ray path, as well as the ray paths associated with all located impacts and shallow moonquakes, we will incorporate the accumulated travel time anomaly as time shifts made to the traces prior to stacking in our array processing technique. This approach will permit a refined seismic constraint on the lunar core, with the side benefit of establishing uncertainty estimates on the model shown in Figure 1.

Modeling deep structure through synthetic seismograms

The simulation of seismic wave propagation through the lunar interior provides direct predictions for the effects of source parameters such as event location and focal mechanism on the travel times and amplitudes of seismic waves. Synthetic seismograms also permit the comparison of observed seismic data to predictions made from differing structural models, serving as an empirical test for the validity of hypothetical depth profiles and permitting the full use of the information contained in each recorded waveform (Figure 2).

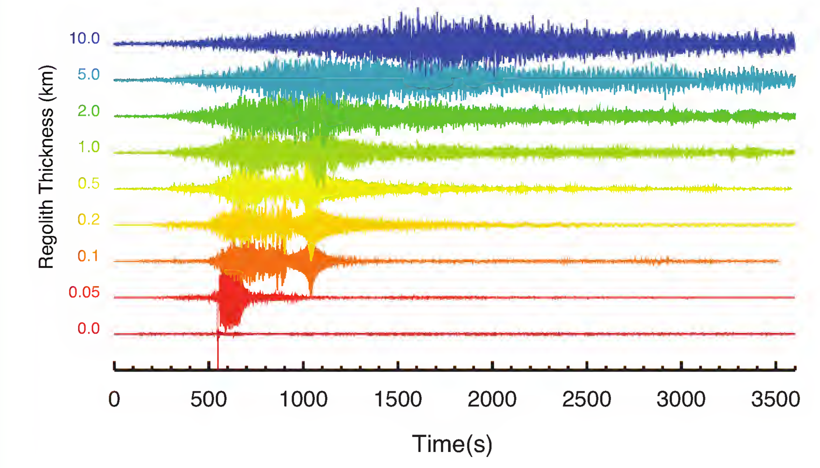

Figure 2: 1-D reflectivity synthetic seismograms for varying thicknesses of the regolith layer (v_p = 1.0 km/s) at the surface of the model (epicentral distance 900 km). A thicker regolith more accurately reproduces the longer rise time observed in real Apollo seismograms.

By inputting synthetic seismograms computed from a given structure model into the array processing procedure, we can iteratively forward model the structure within the Moon to best reproduce our previous observations of deep layer reflections, and test the sensitivity of our results to fine-scale perturbations in the model.

We produce synthetics for the exact moonquake-sensor geometries for the Apollo data. Stacking synthetics following the same procedure for real data enables assessment of the effects of model perturbations on goodness of fit. GRAIL will improve our understanding of the structure of the deep interior, including inferences on core size (fluid and solid parts), possible layering above the core (partial melt) and crustal thickness. Thus our testing will involve perturbing the layer sizes and properties of the best fit structure model shown in Figure 1.

Results for a preliminary test model are shown in Figure 3. For a simple 4-layer model consisting of a solid inner core, fluid outer core, partial melt boundary layer, and constant-velocity mantle (no crust), we computed a suite of synthetic seismograms representative of the known deep moonquake population as recorded at the Apollo stations, and input these traces into our array processing scheme.

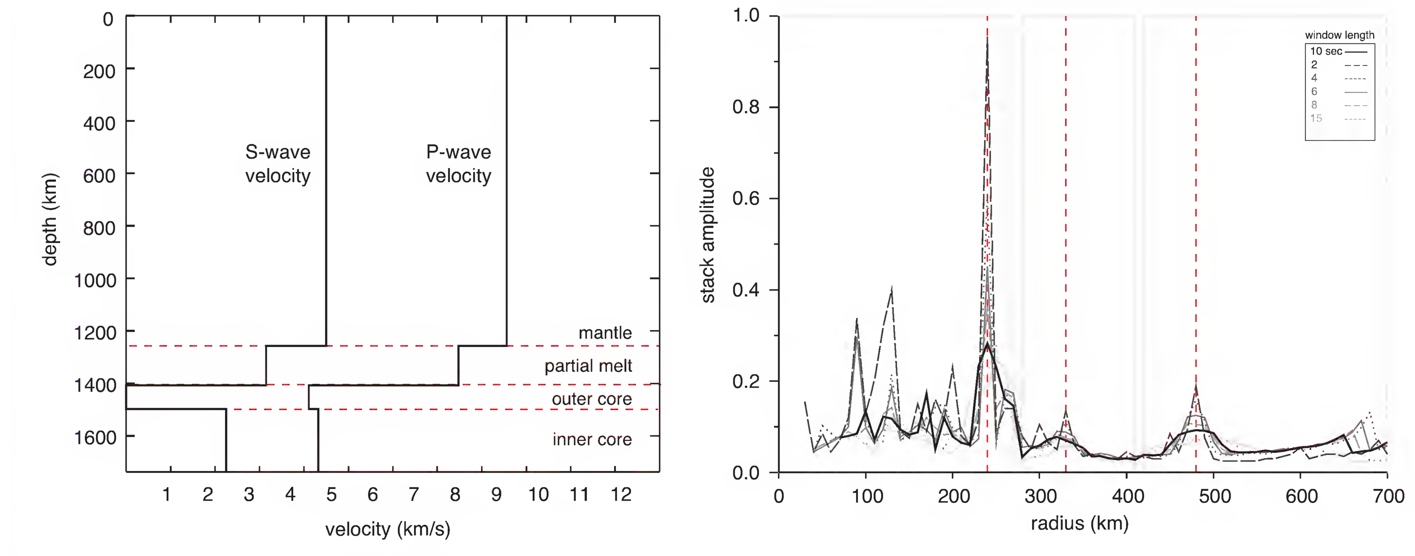

Figure 3: (left) Simplified 4-layer structure model. (right) Array stacking result showing core reflection energy peaks at the appropriate layer radii (dashed red lines).

Joint seismic and gravity inversion

Gravity and seismic data sets are well suited to joint inversion because the complementary information reduces inherent model ambiguity. We will perform a joint inversion [2] of Apollo seismic delay times and GRAIL gravity data in order to recover seismic velocities and density as a function of latitude, longitude, and depth within the Moon.

We relate density (ρ) to seismic velocity (v) using a linear relationship [3]. This relationship is allowed to be depth-dependent, and the corresponding coefficient (B) is only approximately known and can reflect a variety of material properties that vary with depth, including temperature and composition. The inversion seeks to recover the set of density, velocity, and B-coefficient perturbations that minimize (in a least-squares sense) the difference between the observed (d_{obs}) and calculated data (d_{calc}).

	d_{obs}	d_{calc}
seismic data	P- and S-wave arrival times read from recorded seismograms	P- and S-wave arrival times predicted from existing structure model
gravity data	map-projected radial gravity anomaly	a space-dependent scalar estimated point-by-point from the input layer-cake velocity and density model

The model is parameterized using density blocks and velocity nodes. The B-coefficient links density and velocity in each horizontal layer; the vertical density and velocity layer boundaries are required to be common. The lateral and depth extent of the modeled region is dictated by the seismic data coverage, as the GRAIL gravity coverage will be global (Figure 4). Ray coverage from moonquakes does not extend deeper than ~1200km, due to the lack of farside receivers and likely attenuation effects of the core. To prevent edge effects, we will model the entire extent of the nearside, leaving out those nodes that are not pierced by seismic rays.

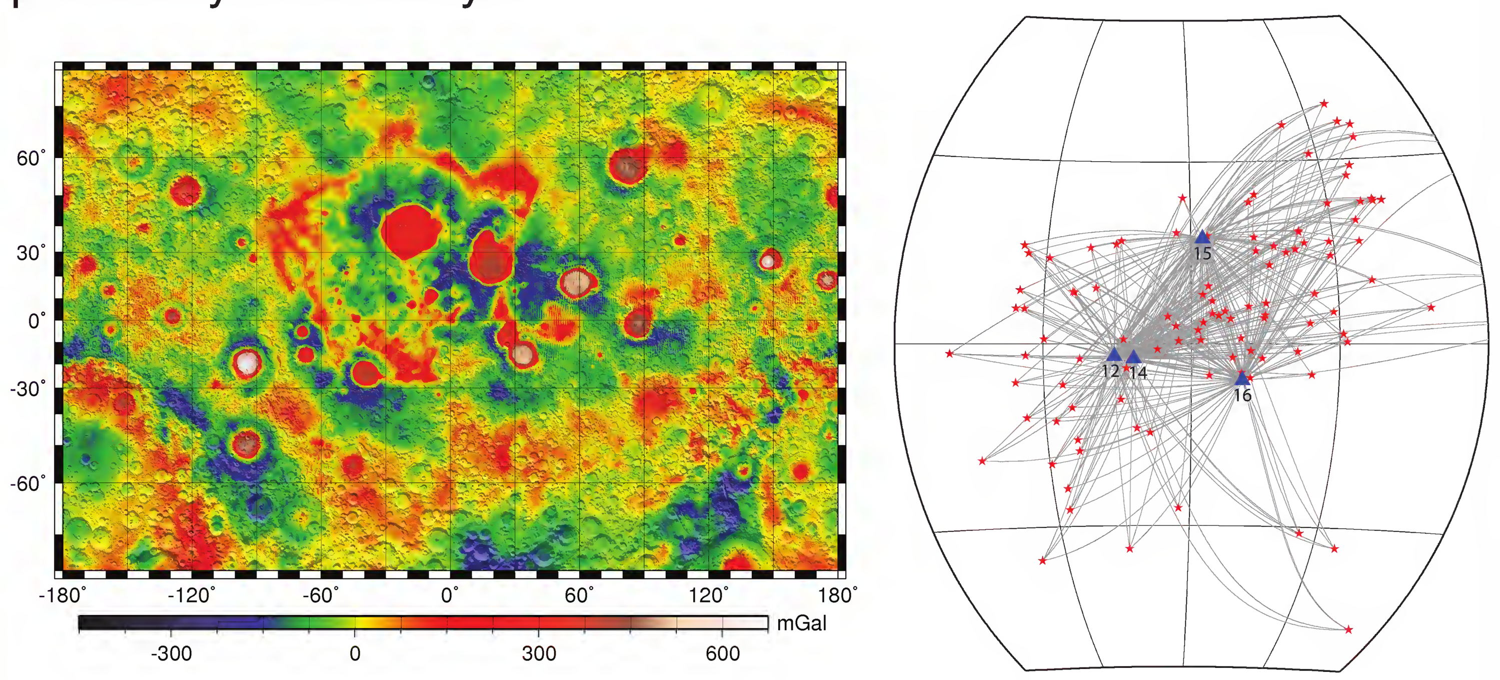

Figure 4: (left) Near-side centered bouger gravity anomaly. (right) Near-side P-wave ray coverage from the deep moonquake population. Red stars mark moonquake epicenters. Blue triangles show the locations of the Apollo seismic stations.

The velocity, density, and B-coefficient perturbations obtained for every layer after each inversion will be applied to the reference model, and the entire process can be repeated iteratively until the root-mean-square misfit stabilizes. This will result in a final model that best fits the constraints jointly imposed by the seismic and gravity observations.

References: **[1]** Weber, R. C.; Lin, P.; Garnero, E. J.; Williams, Q.; Lognonné, P. (2011) Seismic detection of the lunar core, Science 331, 309–312. **[2]** O'Donnell, J. P.; Daly, E.; Tiberi, C.; Bastow, I. D.; O'Reilly, B. M.; Readman, P. W.; Hauser, F. (2011) Lithosphere–asthenosphere interaction beneath Ireland from joint inversion of teleseismic P-wave delay times and GRACE gravity. Geophys. J. Int. 184, 1379–1396. **[3]** Zeyen, H; Achauer, U. (1997) Joint inversion of teleseismic delay times and gravity anomaly data for regional structures: theory and synthetic examples. In Upper mantle heterogeneities from active and passive seismology, Proceedings of the NATO Advanced Research Workshop, Moscow, Russia, 155–168.